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Variability in Gulf Stream Surface-Subsurface Frontal Separation: The Unimportance of Ekman Advection

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Simultaneous observations of the Gulf Stream surface and subsurface (15°C at 200 m) fronts have shown their relative positions to be quite variable. Cross-frontal Ekman advection of the surface front by the local wind stress is a possible major source of this variability. We investigate this possibility and show that on the average and for a particular set of observations, only a small fraction of the observed variability in the Gulf Stream surface-subsurface frontal separation is due to advection of the surface front by the local wind stress. As an alternative, it is suggested that most of the observed variability in frontal separation may be due to unstable meandering of the Gulf Stream surface front.

(Reprints. etc)

1. INTRODUCTION

The only way to monitor large areas of the ocean on a near-real-time basis is to use satellite sensors. However, a problem fundamental to all oceanographic sensors on satellites is that they only sense the ocean's surface, and the subsurface structure of the ocean can only be inferred. While a very few applications may rely upon surface observations such as wave height, the majority of them demand information about the structure of the ocean beneath the surface. In general, the problem of inferring the subsurface structure of the ocean is very broad. However, the fundamental stability of water masses, at least beneath the mixed layer, gives us a way to simplify the problem or at least to reduce its scope. By being able to infer the paths of the subsurface boundaries between water masses, we would be able to account for a substantial part of the ocean's local variability in temperature and salinity.

As an example, the path of the surface front of the Gulf Stream has long been charted. However, many observers have found the surface-subsurface frontal separation of the Gulf Stream to be variable spatially and temporally. The first to examine the Gulf Stream surface-subsurface frontal separation in detail were *Hansen and Maul* [1970]. They found the mean separation between the Stream's surface and subsurface fronts to be 14.5 km with a standard deviation of 12 km. The surface front was found as far as 68 km to the left and 13 km to the right of the subsurface front (looking downstream). *Horton* [1984*a*] observed the surface front as far as 30 km to the right of the subsurface front.

Several different dynamical processes are capable of changing the separation between the Gulf Stream's surface and subsurface fronts. Surface-trapped eddies several tens of kilometers in extent, often having folded waveforms, can change the surface-subsurface frontal separation in their vicinity. For example, *Robinson et al.* [1974] observed that the Gulf Stream's surface front showed a great deal of fine-scale structure, with spatial scales of 30–50 km, that was completely lacking in the path of the 15°C isotherm at 200 m. Observations illustrated in this paper also show these eddies.

Another process commonly observed is the changing width of the Gulf Stream from meander peak to meander trough. As *Newton* [1978] demonstrated, this is because of the changing sign of the Gulf Stream's centripetal acceleration between meander peak and meander trough. The changing width of the Gulf Stream causes a like change in the surface-subsurface frontal separation as demonstrated by *Hansen and Maul* [1970]. They found that on the average, the Gulf Stream's surface-subsurface frontal separation is about 5 km greater at meander peaks than at meander troughs. In the work by *Horton* [1987] additional observations supporting this effect are shown.

A very straightforward way in which the Gulf Stream's frontal separation can vary is through cross-frontal Ekman advection of the surface front by the local wind stress. In the work by *Horton* [1984*b*] the passage of a tropical storm parallel to the Gulf Stream was observed to have apparently increased the surface-subsurface frontal separation by 18 km. However, the passage of a tropical storm is an atypically strong wind stress event. Observations from three surveys of the Gulf Stream are provided which show tens of kilometers changes in frontal separation over 3-day intervals. Using wind velocity measurements from a nearby National Oceanographic and Atmospheric Administration (NOAA) buoy, the Ekman advection of the Gulf Stream's surface front is computed and shown to have been capable of changing the Gulf Stream's surface-subsurface frontal separation by only a few kilometers in the time intervals between the surveys. To complement this calculation, we will show that on the average, the local wind stress is only responsible for a small fraction of the variability in the Gulf Stream's surface-subsurface frontal separation. This will be done by computing the root-mean-square (rms) variability in the frontal separation using a knowledge of the wind stress spectrum.

2. SPECIFIC EXAMPLES OF CHANGES IN FRONTAL SEPARATION

Observations of the Gulf Stream are described which show specific examples of tens of kilometers changes in surface-subsurface frontal separation not due to Ekman advection of the surface front. The observations are three surveys of the

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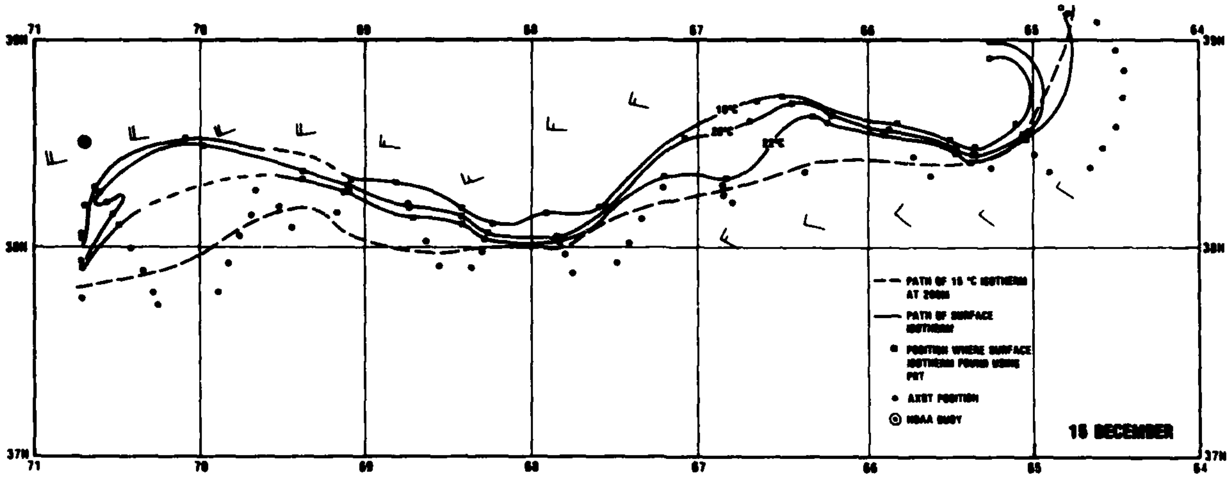

Fig. 1. Plots of surface isotherms encompassing the surface front and the subsurface front during the first flight.

Gulf Stream downstream of Cape Hatteras obtained during December 1982. These surveys were planned to measure the horizontal separation between the paths of the Gulf Stream's fronts at the surface and at 200 m over downstream extents of several hundred kilometers. Assuming the existence of a well-defined mixed layer, the surface front must extend to the base of the mixed layer. Otherwise, there would be a vertical temperature gradient which would contradict the definition of a mixed layer. For this reason we recognize that, in general, the surface front is the surface expression of the mixed-layer front. Therefore when we refer to the subsurface front we really mean the front beneath the mixed layer. As is common, we define the path of the subsurface front to be the path of the 15°C isotherm at 200 m. While this isotherm-depth combination is an excellent indicator of the position of the front of the Gulf Stream, other isotherm-depth pairs could be used, depending upon the depth of interest.

2.1. *Experimental Design*

The observations were conducted using RP-3A flights on December 15, 18, and 21, 1985. Each flight was intended to survey the same zonal segment of the Gulf Stream from 70°40′W to 65°W. However, the second survey flight extended eastward only to 67°W owing to poor weather conditions east of that longitude. Closely spaced air-dropped expendable bathythermographs (AXBT) were used to map the path of the subsurface front. The surface front was mapped primarily by the aircraft's precision radiation thermometer (PRT). Surface temperatures from the AXBTs were used to calibrate the PRT measurements and as an additional source of data where rain made the PRT useless. Rain, thin clouds, and haze were common beneath the aircraft and degraded the reliability of the PRT measurements. Largely because of these atmospheric problems, the uncertainty in the PRT temperatures is about 0.5°C. This uncertainty was estimated by observing the scatter between simultaneous AXBT surface temperatures and PRT temperatures. Additionally, wind speeds at the aircraft altitude (nominally 330 m) were measured using the aircraft's inertial navigation system.

2.2. *Observations*

Figures 1–3 illustrate the surface and subsurface frontal paths obtained during the three surveys. In addition to showing the paths of the 15°C isotherms at 200 m, the figures plot three surface isotherms (18°, 20°, and 22°C) encompassing the region of strong cross-stream surface-temperature gradients which we call the surface front. Because the 20°C isotherm was near the center of the high-gradient region, we will define

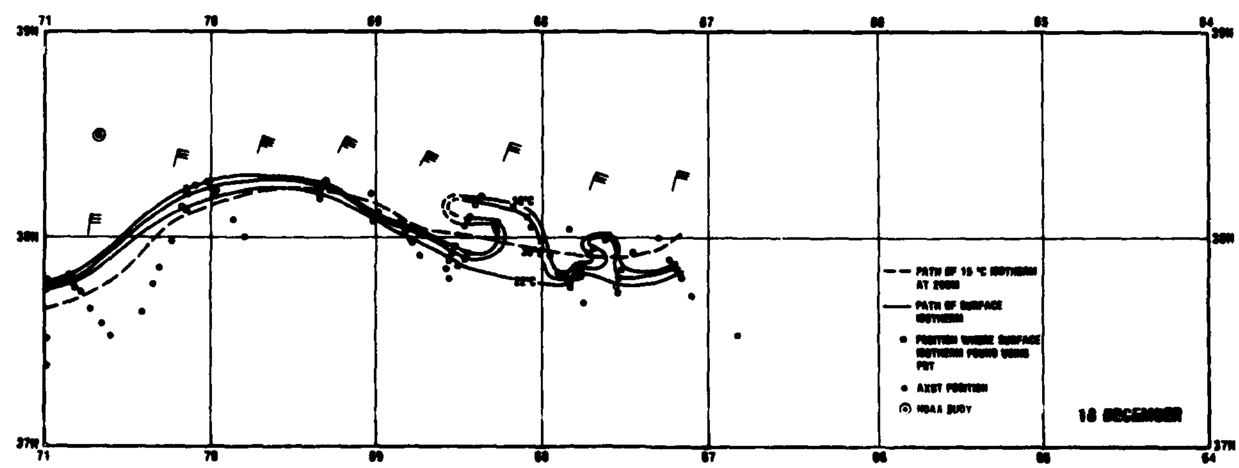

Fig. 2. As in Figure 1, except for the second flight. Surface isotherms are dashed curves where they are uncertain.

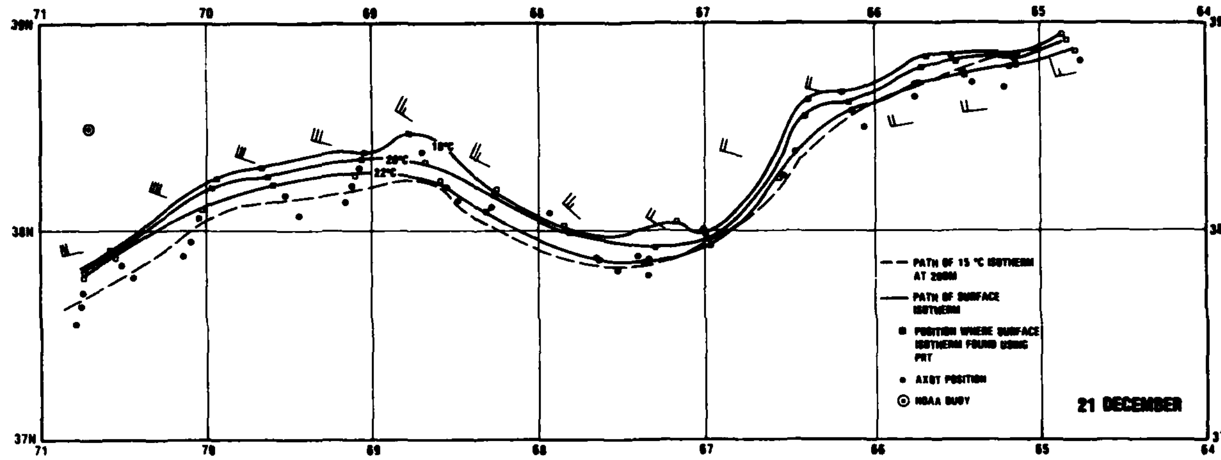

Fig. 3. As in Figure 1, except for the third flight. Surface isotherms are dashed curves where they are uncertain.

the separation between the surface and subsurface fronts as the separation between the 15°C isotherm at 200 m and the 20°C isotherm at the surface.

The winds shown in Figures 1–3 are vector averages over every half degree of longitude. These averages may be compared to the 1600 UT measurements at the nearby NOAA buoy as shown in Figure 4. Multiplying the aircraft-measured wind speeds by 0.8, there is excellent agreement between the buoy winds and the aircraft-measured winds west of 69°W longitude. Wind speed magnitudes agreed within 4 knots, which is the resolution of the wind barbs after the 0.8 correction factor was used. Wind directions agreed within 10°. The apparent 20% reduction in wind speeds from 330 m altitude to those measured at 10 m by the buoy agrees with *Powell* [1980] and is indicative of an unstable atmospheric boundary layer. A greater reduction in wind speed would be expected in stable conditions.

During the first survey on December 15 the separation between the surface and subsurface fronts was quite variable. The variability was due to an approximately 300-km wavelength meander, the amplitude of which was much greater at the surface than at 200 m. The greatest separation of 60 km occurred at the peak of the meander near 70°W, while the minimum separation of a few kilometers occurred at the troughs of the meander near 68° and 65°W. The changes in separation may initially be interpreted as a very strong dependence of the separation between the surface and subsurface fronts upon curvature of the subsurface front or that there is a surface-trapped meander. However, if the change in frontal separation from peak to trough is due to the curvature effect,

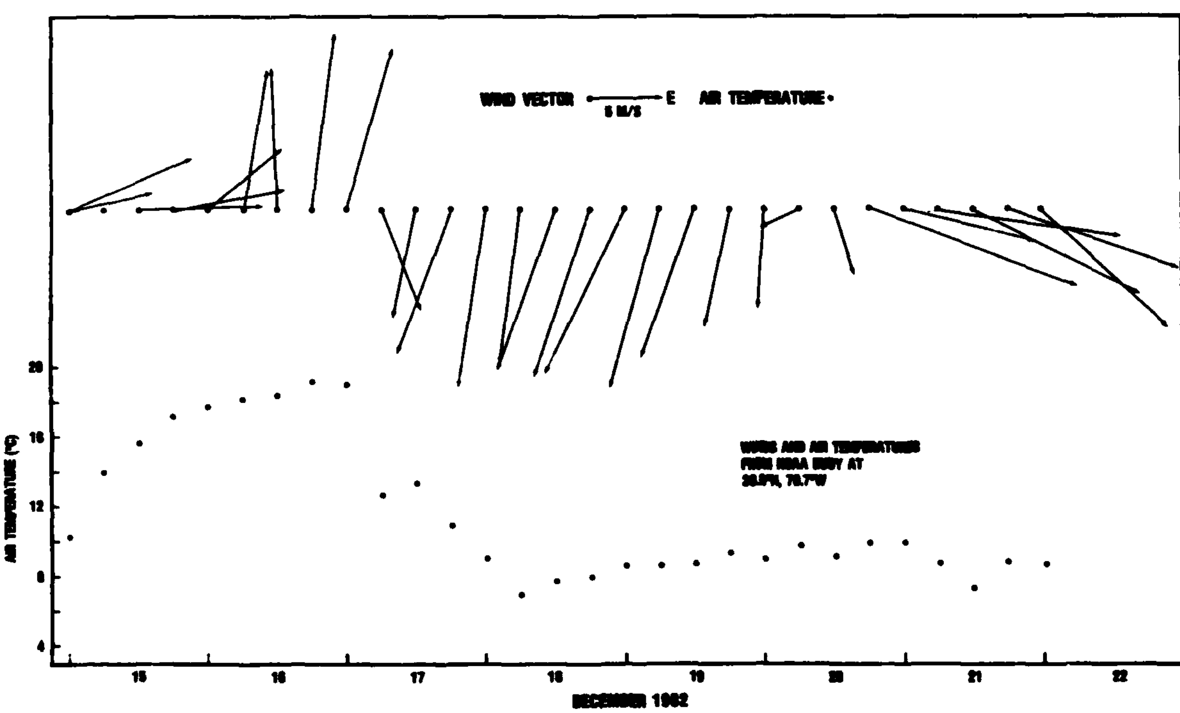

Fig. 4. Wind vectors and air temperatures measured at a NOAA buoy north of the stream during the field experiment. Buoy position is marked on Figures 1–3.

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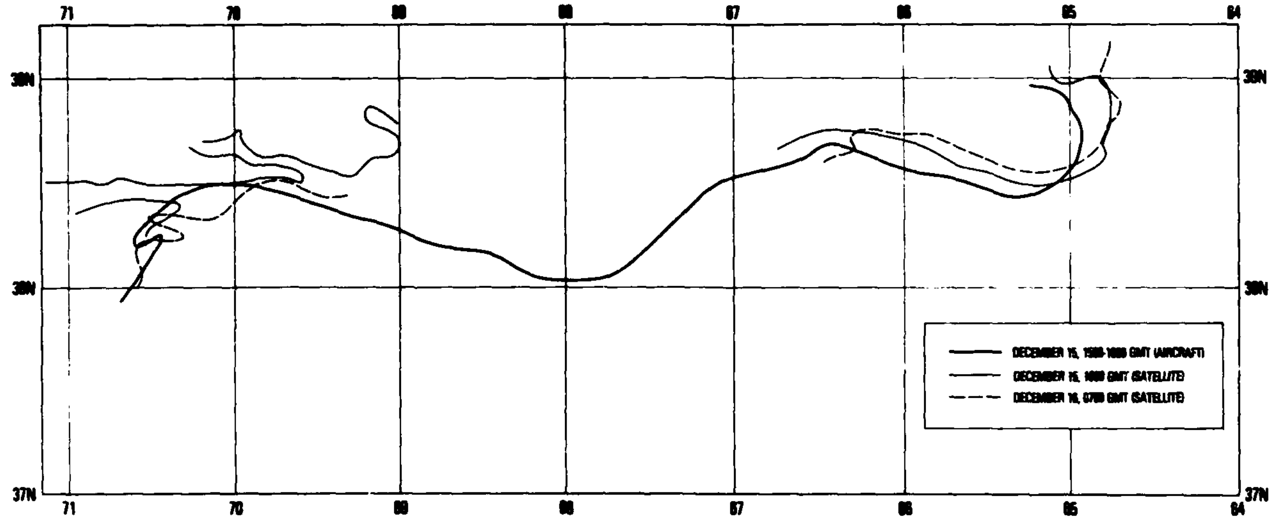

Fig. 5. Comparison between gulf stream surface front (20°C) paths determined from aircraft PRT measurements and satellite IR imagery.

the curvature effect is several times stronger than predicted by the theory of *Horton* [1987]; this is because of the small 10–15 km) amplitude of the meander in the subsurface front.

Figure 5 shows paths of the Gulf Stream's 20°C surface isotherms obtained from two NOAA 8 multichannel infrared images within a day following the first survey flight. The images were corrected for water vapor resulting in less than a degree uncertainty in computed temperatures. Land was not visible in the imagery, and as a result the positioning is not ground control point matched. Also shown is the path the 20°C isotherm obtained from the first flight; reasonable agreement is shown. Differences are mainly systematic and are probably due to (unknown) positioning errors in the satellite IR image and the uncertainties in the surface temperatures from the satellite image and our aircraft's radiation thermometer. Most interesting is the apparent rapid variability in the shape of the folded wave or spin-off features visible in the satellite imagery.

Comparing Figures 1 and 2 we see that in general, the frontal separation was quite different during the secondary survey on December 18 than during the first survey on December 15. Near 70°W the frontal separation decreased about 50 km between the first and second surveys from about 60 to 10 km. The decrease was approximately equally due to a southward movement of the 20°C surface isotherm and a northward movement of the 15°C isotherm at 200 m. Toward the eastern edge of the area covered by the second survey, there was also a decrease in separation between the fronts. However, there the separation was confused by the presence of two folded-wave eddies. Near these folded-wave eddies the 20°C isotherm was approximately 10 km to the right (looking downstream) of the subsurface front so that there the change in frontal separation was about 20 km. The subsurface front moved little, and the change was basically due to a southward movement of the surface front. As is normal for folded-wave eddies, they did not manifest themselves in the path of the 15°C isotherm at 200 m, and they gave the impression of being sheared by a differential amount of downstream advection.

The third survey on December 21 covered the same zonal segment of the Gulf Stream as did the first survey. The surface and subsurface fronts tracked closely and, overall, the separation between the surface and subsurface fronts was intermediate between that of the first and second surveys. Near 68°W longitude during the third survey, the surface front was about 10 km north of the subsurface front. There the surface front moved roughly 20 km relative to the subsurface front during the 3 days between the second and third survey flights. Moving west of that longitude, the change in frontal separation between the second and third surveys decreased until at the extreme western edge of the survey area there was no change. While the gap in the sampling of the surface temperatures near 67°W created some uncertainty, the folded-wave eddies were gone. The variability in the width of the surface front between 68° and 66°W may be their remnants. The possible rapid evolution of the folded-wave eddies is consistent with *Stern* [1985]. The amplitude of the approximately 300-km wavelength meander in the subsurface front did not increase between the second and third surveys, but there was a downstream phase shift of the meander. Based upon the movement of the meander crest, the downstream phase speed was about 30 cm/s. This estimate is greater than the phase speeds of comparable meanders observed by *Robinson et al.* [1974] and by *Hansen* [1970]; these estimates were 20 and 5–10 cm/s, respectively.

2.3. *Discussion*

The observations show changes in frontal separation of up to several tens of kilometers over 3-day intervals between the three surveys. Several processes could, in principle, have been responsible for the changes in frontal separation. For example, the curvature effect causes, on the average, the frontal separation to be less at meander troughs than at meander peaks. However, because tens of kilometers changes in frontal separation occurred over most of the survey region and not just at peaks and troughs, other processes capable of changing frontal separation must be important. Another possibility that can be discounted is that a change in frontal separation of unknown origin propagated from the west into the survey region.

An example of the importance of downstream movement of changes in frontal separation is shown in the Gulf Stream

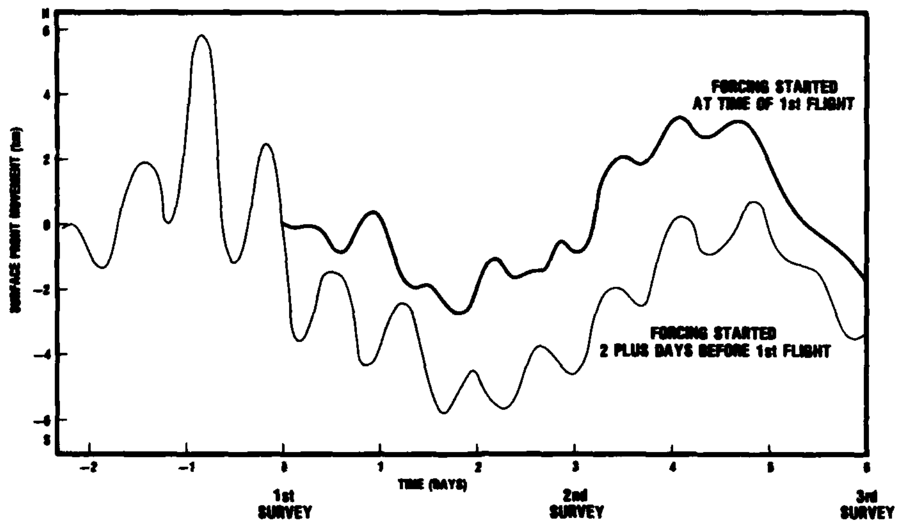

Fig. 6. Predicted displacement of the surface front due to wind-driven mixed-layer advection.

surveys of *Horton* [1984*b*]. In those surveys an eddy or folded-wave feature was observed in the surface front. Downstream of this eddy the Gulf Stream's surface front was to the left (looking downstream) of the subsurface front while upstream the surface front was to the right of the subsurface front. This eddy which separated regions of dissimilar frontal separation moved downstream between survey flights with a speed of 50 cm/s. In the present survey substantial changes in frontal separation occurred, in general, over most of the survey area and did not appear to originate from the inflow side of the survey area. Except for the shortened second survey flight, the survey area covered 550 km of the Gulf Stream's path. Thus a change in frontal separation would have to move with a downstream speed in excess of 210 cm/s to cross the entire survey area between flights. Furthermore, at the extreme western edge of the survey area, the surface-subsurface frontal separation did not change by more than a few kilometers between survey flights.

In *Horton* [1984*b*] we modeled cross-frontal Ekman advection of the surface or mixed-layer front due to the passage of a tropical storm. In this case the storm passage apparently was responsible for an 18-km change in frontal separation. Using a model very similar to that described in the work by *Horton* [1984*a*], we modeled cross-frontal advection of the surface front by the local wind stress during the survey period.

Because of the excellent agreement between the aircraft-measured winds west of 69°W and the winds measured at the NOAA buoy at 38.5°N and 70.7°W, we used the buoy winds to compute wind stress. Wind speed, wind direction, and air temperature were obtained from the buoy at 6-hour intervals. These measurements were linearly interpolated to provide continuously variable values. As was noted, the change in the separation between the surface and subsurface fronts varied with downstream position. Because the NOAA buoy was near the western side of the survey area, this is where the modeled change in separation should be compared to the observed change. However, temporal changes in the strength of the curvature effect where the flow curvature was anticyclonic at the western edge of the survey area may have contributed to the observed changes in the separation. For this reason the modeled changes in separation should be compared to the observed changes at the westernmost inflection point for the flow curvature near 69°W longitude.

The model was initialized with a mixed-layer depth of 70 m representing the average mixed-layer depth in the frontal region. Mixed-layer depth was allowed to change via Ekman pumping and entrainment; the *Niiler and Kraus* [1977] entrainment model was used. However, winds were not strong enough to do more than a trivial amount of entrainment during the modeling period. This is consistent with observations which show no significant entrainment or detrainment during the modeling period.

Model results are shown in Figure 6. Initially, the model was run with the winds starting when the first survey was taken and assuming zero cross-stream flow in the mixed layer. The modeled movement of the surface front is minimal, especially in comparison to the observed displacement of the surface front almost anywhere in the survey area. The predicted displacement of the surface front was never more than a few kilometers and at the time of the second and third surveys was within a kilometer of its original position.

Since the *e*-folding decay time for near-surface inertial oscillations is a few days, mixed-layer current speeds depend upon the history of the wind stress over the past few days. Because of this several model runs were made starting the wind stress ahead of the first survey. The results of the model run showing the greatest displacement of the surface front are also illustrated in Figure 6. Defining the displacement of the surface front to be zero at the time of the first survey, the displacement of the surface front was 4+ km to the south at the time of the second survey and 3 km to the south at the time of the third survey. Still, this response is small compared to the tens of kilometers changes in separation observed.

3. Root-Mean-Square Variability in Frontal Separation Due to Ekman Advection

Given a knowledge of the wind stress spectrum, we can estimate the rms variability of the Gulf Stream's surface-subsurface frontal separation. What is actually computed is the rms variability of the cross-stream position of a parcel in

the mixed layer. Again, we assume that the surface front is the surface expression of the mixed-layer front. In addition, we assume that the velocity profile in the mixed layer is slablike or independent of depth. Consistent with these assumptions, the rms variability in the cross-stream position of a parcel in the mixed layer will produce an equal rms variability in the Gulf Stream's surface-subsurface frontal separation.

In filtering theory [*Bloomfield*, 1976] a common relationship is

$$S_R(\omega) = |\hat{g}|^2 S_f(\omega) \quad (1)$$

where S_R and S_f are the response and forcing spectrums, respectively, and ω is frequency. The transfer function $\hat{g}$ is the Fourier transform of the impulse response g. An example of the use of (1) in the oceanographic literature is found in the work by *Kundu* [1984].

In order to obtain the response (the transfer function $\hat{g}$) to a forcing spectrum, we assume that linear dynamics govern the current speeds in the mixed layer. For simplicity, we assume a meridional current with a mixed layer of constant depth h. If u and v are the zonal and meridional speeds in the mixed layer, f is the Coriolis parameter, k is a damping coefficient, and τ is the meridional wind stress,

$$u_t - f(v - V_0) = -ku \quad (2)$$

$$(v - V_0)_t + fu = -k(v - V_0) + \tau/h \quad (3)$$

The geostrophic component of the current speed, assumed to be steady, is V_0. In (2) the geostrophic balance has been subtracted. In forming (3) it was assumed that kV_0 balances the downstream pressure gradient. Equations (2) and (3) can be combined to give

$$X_{ttt} + 2kX_{tt} + (k^2 + f^2)X_t = f\tau/h \quad (4)$$

where $X_t = u$. This equation describes the cross-stream displacement X of a water parcel in the mixed layer. We shall use this equation to determine the impulse response $g(t)$ to a unit impulse forcing of the mixed layer by the meridional wind stress. Taking the Laplace transform L of (4) gives

$$L(X) = \frac{1}{P(S)} L(f\tau/h) \quad (5)$$

where

$$P(S) = S^3 + 2kS^2 + (k^2 + f^2)S \quad (6)$$

The impulse response [*Kreider et al.*, 1966] is

$$g(t) = L^{-1}(1/P(S)) \quad (7)$$

This impulse response assumes that the cross-stream displacement X and all of its temporal derivatives are zero at $t = 0$. Using (6) and (7), it is straightforward to show that

$$g(t) = \frac{f - ke^{kt} \sin ft - fe^{-kt} \cos ft}{f(k^2 + f^2)} \quad (8)$$

The impulse response g is equal to zero at $t = 0$ and is presumed to be zero when $t < 0$.

From (4) it is apparent that the forcing spectrum in (1) is $f^2 S_\tau / h^2$, where S_τ is the wind stress spectrum. For our case the response spectrum S_R in (1) is the spectrum of the cross-stream displacement X of a parcel in the mixed layer. Using (1) and because $\int_0^\infty S_R \, d\omega$ is $\overline{X^2}$, the mean-square cross-stream displacement, the relationship between S_τ, $\hat{g}(\omega)$, and $\overline{X^2}$ is

$$\overline{X^2} = \int_0^\infty f^2 \frac{S_\tau}{h^2} |\hat{g}|^2 \, d\omega \quad (9)$$

Equation (8) shows that part of $g(t)$ is a constant. The Fourier transform of a constant will yield a delta function centered at zero frequency. Thus part of $|\hat{g}|^2$ will be a constant multiplied by the delta function squared. Because the integral of the square of the delta function is infinite, a S_τ which is nonzero at zero frequency will yield an infinite mean-square cross-stream displacement of a water parcel in the mixed layer. This makes sense because a steady wind parallel to the current will drive a constant mixed-layer flow normal to the current. An arbitrarily large cross-stream displacement of a parcel in the mixed layer would result. We do not allow a steady wind and will accordingly assume that $S_\tau(0) = 0$. Because of this assumption we can ignore that part of $g(t)$ which is a constant and instead use

$$g(t) = \frac{-ke^{-kt} \sin ft - fe^{-kt} \cos ft}{f(k^2 + f^2)} \quad (10)$$

Using this definition for $g(t)$

$$\hat{g}(\omega) = -(k + if)(f - \omega - ik)[2f(k^2 + f^2)(k^2 + (f - \omega)^2)]^{-1}$$
$$-(k - if)(f + \omega + ik)[2f(k^2 + f^2)(k^2 + (f + \omega)^2)]^{-1} \quad (11)$$

As expected, $\hat{g}(f)$ is infinite if there is no damping of inertial oscillations ($k = 0$). The resonant peak in the transfer function is approximately proportional to $1/k$.

We estimated the root-mean-square cross-frontal displacement of the surface front using a downstream wind speed–squared spectrum constructed using hourly downstream winds taken between December 1983 and February 1984 at the NOAA buoy north of the Gulf Stream at 70.5°W longitude. The downstream winds were assumed to be the component of the wind velocities parallel to the mean path of the Gulf Stream at the longitude of the NOAA buoy. The spectrum shown in Figure 7 is not plotted for periods greater than 40 hours because the modeled transfer function is insignificantly small at longer periods. As a check on this spectrum, we estimated a second wind speed–squared spectrum from a variance-conserving wind speed spectrum obtained 20 km offshore of Savannah by *Schwing and Blanton* [1984] during the early summer of 1977. This was done by squaring their spectrum and then dividing by frequency giving a wind speed–squared spectrum. This spectrum is the total wind speed–squared spectrum. The downstream wind speed–squared spectrum was estimated by dividing the total wind speed–squared spectrum by two.

Multiplying a wind speed–squared spectrum by $(C_D \rho^*)^2$ gives the S_τ. C_D is the drag coefficient and ρ^* is the air density divided by the water density. $C_D \rho^*$ was assumed to be 1.6 $\times 10^{-6}$. As *Huang et al.* [1986] discuss, drag coefficients are probably sensitive to surface wave roughness. Because of the Gulf Stream's current shear, surface wave roughness in the vicinity of the Gulf Stream is strongly modified. Based upon the scatter in the C_D values tabulated by *Huang et al.* [1986], our assumed $C_D \rho^*$ may be in error by up to a factor of 2. As is shown by (9), the forcing term is the wind stress divided by the mixed-layer depth. When using the December–February and summer wind stress spectrums, we assume mixed-layer depths of 85 and 30 m, respectively. On the basis of mixed-layer depth alone, strong seasonal changes in the predicted cross-frontal displacement should be expected. Because the diameter of an inertial circle is inversely proportional to the magnitude

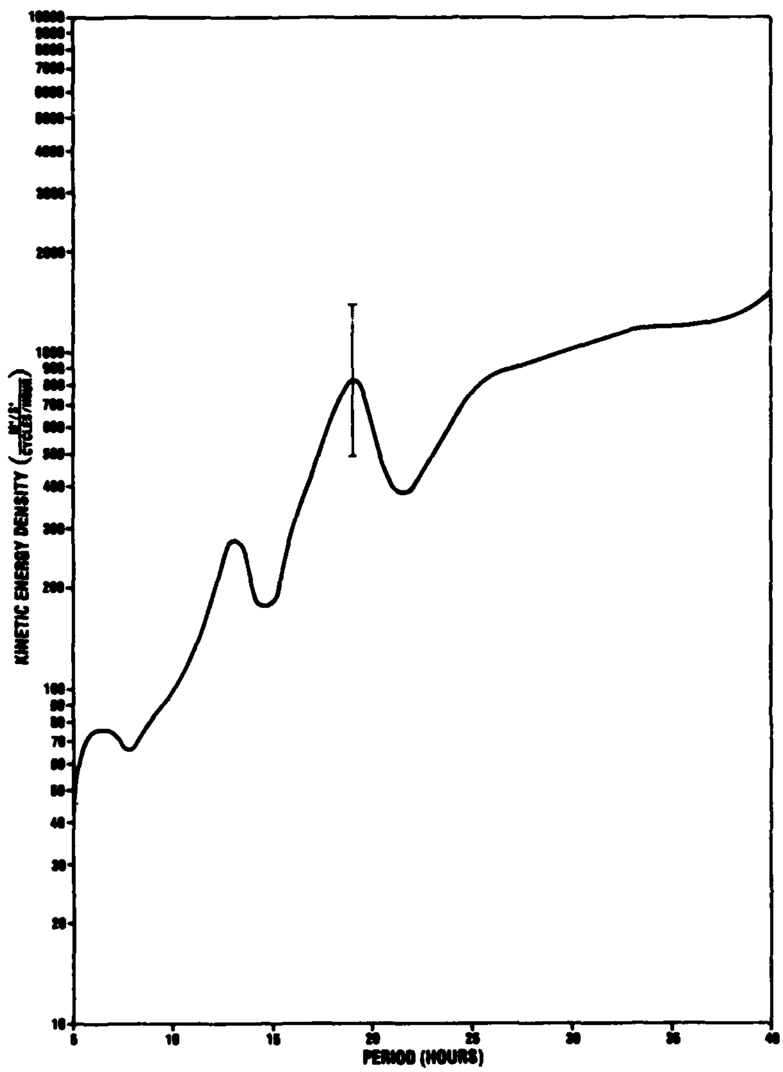

Fig. 7. Downstream wind speed-squared spectrum used to predict the rms variability of the stream's surface-subsurface frontal separation.

of f, another important but smaller factor affecting the magnitude of predicted rms displacement is latitude. With everything else being equal, the predicted rms displacement for the Gulf Stream near the NOAA buoy at 38.5°N would be about 20% smaller than the predicted displacement at the latitude of Savannah.

Since $\hat{g}(\omega)$ and S_τ are now known, $\overline{X^2}$ can be computed using (9). Because of the complexity of $|\hat{g}|^2$, the integral in (9) was evaluated numerically. The S_τ was assumed to be zero at frequencies lower and higher than those given values in Figure 7. During the integration the mixed-layer depth was assumed to be 30 m, and the damping constant k was assumed to be 1/5 days. This value is consistent with *Pollard and Millard* [1970], who found that modeling observed wind-generated inertial oscillation required 2–8 day *e*-folding damping times.

Using the December–February spectrum the predicted rms displacement $(\overline{X^2})^{1/2}$ was 0.3 km. Near the inertial frequency wind speeds were considerably higher in December than during the following January and February. Using December winds alone, the predicted rms displacement was 0.6 km. In contrast, the summer spectrum yielded a greater rms displacement of 1.3 km. While the summer wind stress spectrum may have been less accurate because it was inferred from a wind speed spectrum, the summer spectrum would have yielded about 3 times the rms cross-stream displacement of the December–February spectrum just on the basis of the differences in latitude and assumed mixed-layer depth alone. The predictions were also sensitive to the frictional coefficient k. Increasing k from (5 days)$^{-1}$ to (3 days)$^{-1}$ decreased the estimate predicted using the December–February spectrum by half while decreasing k to (7 days)$^{-1}$ increased the predicted displacement to 0.4 km.

The modification of the inertial frequency of a mixed-layer parcel by the horizontal velocity shear of the Gulf Stream might quantitatively change our result. To show this we replace fu in (3) with $(f + V_{0_x})u$, where V_{0_x} is the cross-stream shear in the downstream geostrophically balanced current. Doing this we find that f in the left-hand side of (4) and hence in (8), (10), and (11) becomes $f^{1/2}(f + V_{0_x})^{1/2}$. We call this modified Coriolis parameter f^*.

Because of the sensitivity of the transfer function to f^*, current shear alters the overall response of the transfer function. Most important, however, is the shift in the inertial frequency and hence of the frequency where the response of the

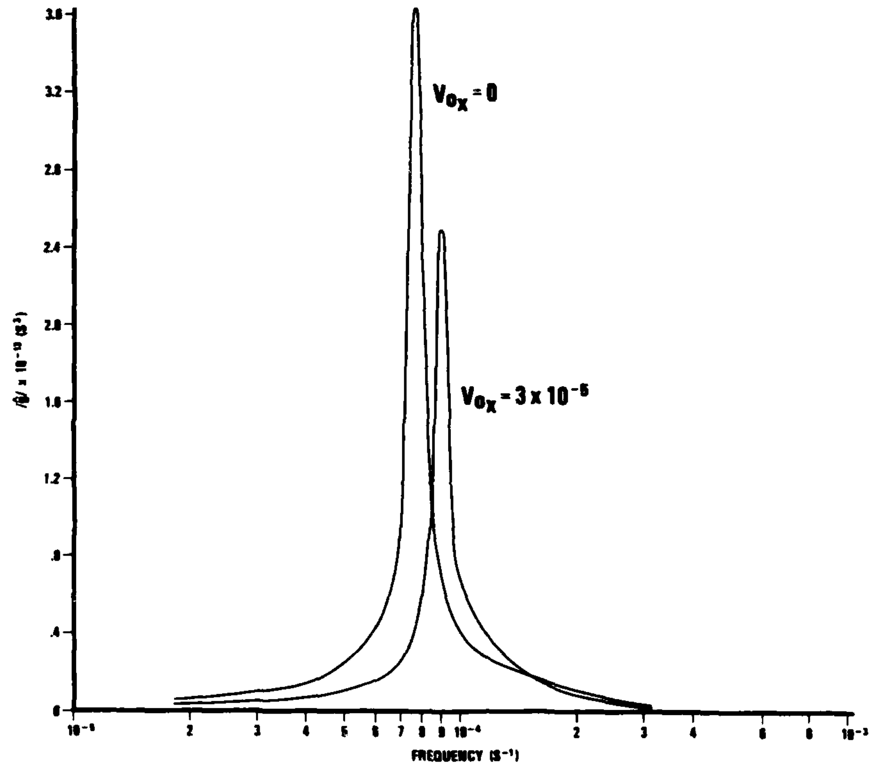

Fig. 8. Transfer functions assuming $f = 0.77 \times 10^{-4}$.

transfer function peaks. Figure 8 illustrates the transfer function for the cases when V_{0_x} is zero and when V_{0_x} is 3×10^{-5} s^{-1}. The positive V_{0_x} is consistent with the inshore edge of the Gulf Stream. Defining f^* assuming V_{0_x} to be 3×10^{-5} s^{-1}, the rms cross-stream displacement inferred from the December–February spectrum is reduced by half. The reduction in the predicted displacement occurred because the modification of f^* by the shear meant that the peak in the transfer function at $\omega = f^*$ no longer coincided with the inertial peak in the wind stress spectrum.

Because of seasonal changes in wind speed and mixed-layer depth we can not make a precise estimate of the rms wind-driven cross-stream displacement. Additionally, there is a factor of 2 uncertainty in the drag coefficient we use to compute wind stress. Equivalent uncertainties are caused by the sensitivities of the predicted solution to V_{0_x} and k. The important result is that the estimates of wind-driven variability are only about 10% or less of the rms variability in the surface-subsurface frontal separation observed by *Hansen and Maul* [1970]. We can reliably state that most of the observed changes in frontal separation are not due to the local wind stress.

4. ALTERNATIVE SOURCE OF VARIABILITY IN FRONTAL SEPARATION

Because the majority of the rms variability in the Gulf Stream surface-subsurface frontal separation cannot be attributed to cross-frontal Ekman advection, there must be an alternative source of variability in frontal separation. Variability in the Gulf Stream surface-subsurface frontal separation can be a result of changes in the cross-sectional shape of the Gulf Stream beneath the mixed layer. For example, between meander peak and meander trough the cross-frontal slope or tilt of the Gulf Stream front changes in response to the reversal in sign of the path curvature and centripetal acceleration of the downstream flow. While this curvature effect can at times be significant as discussed, the effect was not consistent with the tens of kilometers changes in frontal separation described in section 2. More general changes in the cross-sectional shape of the Gulf Stream may be a source of variability in the frontal separation. However, *Halkin and Rossby* [1985] observed that the structure of the Gulf Stream remains remarkably constant and indifferent to its immediate position.

Our observations of changes in frontal separation are consistent with unstable surface-trapped waves or meanders. These would induce greater lateral displacements in the mixed layer than at 200 m. The model of frontal instabilities by *Orlanski* [1968] may be applicable. The system he analyzes contains two homogeneous fluids bounded above and below by rigid horizontal planes. The motion of each fluid, the boundaries of which form the front, is independent of vertical coordinate Orlanski identifies several types of instabilities. The particle motion of one of them (his Rayleigh-like instability *R*) is nearly horizontal and would give rise to cross-frontal displacements. The growth rate for this instability increases with increasing horizontal shear and decreasing Richardson number. *Ramp et al.* [1983] note that the phase speed of instability *R* is to a good approximation that of a Rayleigh shear wave when the wavelength is large compared to the width of the front. Because the surface front is generally only a few kilometers wide, the approximation will be good for any wavelength our field experiment had the ability to resolve. Strictly, Rayleigh shear waves exist on a vertical front for which there is no cross-frontal density gradient. The phase speed of Rayleigh shear waves is the average of the current speeds on both sides of the front. For the Gulf Stream this

puts an upper limit of about 100 cm/s assuming 0 cm/s on one side of the front and 200 cm/s on the other side. The actual observed phase speed should probably depend on the cross-stream position of the surface front relative to the subsurface front. The highest phase speeds would be expected when the surface front is anomalously to the right (looking downstream) of the subsurface front and near the high speed core of the Gulf Stream.

If hydrodynamically unstable waves or meanders of the Gulf Stream's surface front are to cause variability in the Gulf Stream's surface-subsurface frontal separation, the instabilities must be surface trapped. The appropriate parameter governing the vertical scale of penetration H of a quasigeostrophic feature in a stratified fluid is fL/N, where N is the Brunt-Vaisala frequency and L is the horizontal scale over which the feature changes. The scale follows immediately from the quasigeostrophic potential vorticity equation $V_{yy} + (f^2/N^2)V_{zz} = 0$ if we let $V_{yy} = V/L^2$ and $V_{zz} = V/H^2$, where V is the downstream current speed. If instead we assume semigeostrophic dynamics, the scale consistent with the L is transformed to $L + V/f$ [*Hoskins and Bretherton*, 1972]. However, because V/f is only about 10 km, the scale depth of penetration will not be radically different assuming quasi- or semigeostrophic dynamics. Consistent with the observations of *Robinson et al.* [1974], the parameter shows that features should become increasingly surface trapped as the horizontal scale over which they change decreases. This concept has been used by, among many others, *Rhines* [1970] in a study of planetary topographic waves and more recently by *Orlanski and Polinsky* [1983] in a study of ocean response to mesoscale atmospheric forcing. In the Gulf Stream the magnitude of this scale length is established by the width of the subsurface front rather than by how far the surface front moves or by the horizontal extent or wavelength of the meander of the surface front. *Horton* [1984*b*] tested this idea. There, using the scale L determined by the width of the subsurface front, about 17 km, we were able to describe properly the decay with depth h of an eddy in the Gulf Stream's surface front.

5. CONCLUSION

We were able to demonstrate that on the average, Ekman advection of the Gulf Stream's surface front driven by the local wind stress does not contribute significantly to variability in the surface-subsurface frontal separation. Our estimate of the rms variability in frontal separation due to Ekman advection is only about 10% or less of the variability deduced by *Hansen and Maul* [1970] from observations. Additionally, specific observations of tens of kilometers changes in Gulf Stream frontal separation over 3-day intervals are sown. Using wind velocities from a nearby NOAA buoy, we showed that the observed changes in frontal separation were not as a whole due to Ekman advection.

Because of the apparent unimportance of Ekman advection, the dominant source of variability in the frontal separation may be due to unstable meandering of the Gulf Stream's surface or mixed-layer front. It is suggested that this unstable meandering can be strongly surface trapped and hence decoupled from the Stream's front at a 200-m depth because of the narrowness of the Gulf Stream's front.

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